Tide gauge records of uplift along the northern Pacific-North American plate boundary, 1937 to 2001

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Vertical crustal motions at 15 sites along the northern Pacific-North America plate boundary are determined using relative sea level changes from tide gauge records. Our analysis is based on monthly mean sea levels, from which barometric pressure and seasonal effects are removed. The records are corrected for common-mode oceanographic variations. These records are statistically examined for non-linear behavior related to glacial isostatic, tectonic and postseismic effects. To estimate land uplift rates, the local effect of global sea level rise is removed from the relative sea level rates. Slow rates of vertical motion are observed along the southern strike-slip plate boundary. The extremely rapid uplift of the northern strike-slip boundary can be attributed entirely to viscoelastic postglacial rebound associated with tidewater-glacier retreat in Glacier Bay and regional post-Little Ice Age deglaciation. Isostatic modeling indicates a mantle viscosity of $\approx 2 \times 10^{19} - 5 \times 10^{19}$ Pa s, similar to that found elsewhere along the Pacific-North America plate margin. At Yakutat, near the transition of plate motion from strike-slip to subduction, complex non-linear behavior is evident, with a significant change in uplift rate following the 1979 St. Elias earthquake. Non-linear uplift rates are predominant within the 1964 Great Alaskan earthquake near-field. Rapid uplift at Kodiak during a 3.5 year period starting mid-1964 totaled 47 $\pm$ 8 cm. Anchorage, Seward and Seldovia exhibited oscillatory uplift in the period immediately following the earthquake until mid-1972. Since mid-1972, uplift rates have increased steadily at Anchorage, Seward, Cordova, and Valdez. During this period Nikiski and Kodiak show decreasing uplift rates.

INDEX TERMS: 1255 Geodesy and Gravity: Tides—ocean (4560); 1294 Geodesy and Gravity: Instruments and techniques; 4556 Oceanography: Physical: Sea level variations;
KEYWORDS: sea level, glacial-isostasy, uplift, postseismic, Alaska tectonics, tide gauge


1. Introduction

[2] Water levels recorded at tide gauges over several decades or more allow the use of mean sea level (MSL) as a datum against which long-term changes of land level can be compared. Vertical crustal motions can be measured by changes in MSL; if the long-term MSL of a site appears to fall, the land level is rising, and vice versa. Sea level observations allow continuous observation of solid earth motions over a much longer time frame than any other direct measurements (GPS, VLBI, strain gauges, EDM networks, etc.), and thus can provide determinations of long-term and time-varying earth responses to earthquake cycles and changes in surface loading. We present results for vertical crustal motion at 15 permanent tide gauge stations along the northern section of the Pacific Plate boundary with North America, an extremely active tectonic and glacio-isostatic environment (Figure 1).

[3] The tide gauge stations in this region reside in a wide range of tectonic environments, varying from pure strike-slip motion to oblique convergence, and shallow to moderately dipping oceanic/continental subduction (Figure 1). The records from these stations show large coseismic datum shifts [Plafker, 1971], evidence of postseismic relaxation [Savage and Plafker, 1991; Cohen and Freymueller, 2001], and some of the fastest uplift rates in the world [Hicks and Shofnos, 1965]. The wide range of tectonic environments along the northern Pacific Plate boundary, the rapid uplift of many of the sites, and the large earthquakes that many of the records span all provide a unique opportunity to apply MSL analysis to complex and time-varying vertical crustal motion.

[4] One feature common to past tide gauge studies of crustal motion in the Gulf of Alaska and the northeast Pacific is the use of annual MSL data. Because of occasional data gaps, annual MSL is often calculated from an incomplete set of monthly MSL’s, which is problematic [Cohen and Freymueller, 2001]. This region of the Pacific Ocean has the largest seasonal variation in monthly MSL of anywhere in the world ($\approx 50$ cm) [Pugh, 1987], and the absence of even one monthly data point can lead to strong biasing of annual MSL averages.
Figure 1. (a) Map of the study area, showing tide gauge locations. (b) Tectonic and glacial setting. The open arrow is the NUVEL-1A velocity of the Pacific Plate relative to North America [DeMets et al., 1994]. Dark gray areas indicate glaciers and ice fields. Strike-slip motion occurs along the Fairweather-Queen Charlotte fault (FW-QC), while the Pacific plate is subducted beneath North America along the Aleutian Megathrust (AMT). The plate interface that was ruptured from Prince William Sound (PWS) to Kodiak Island during the 1964 Great Alaska Earthquake is outlined with a dashed line. At the transition from strike-slip to subduction, a micro-continent (the Yakutat Block) is actively colliding with North America. The western portion of the leading edge of the Yakutat Block is partially being subducted in the Kayak Island and Pamplona Zones (KI-PZ), while crustal shortening accommodates relative plate motion at the eastern corner of the leading edge in the St. Elias Mts. The Yakutat Block is bounded to the south by the Transition Zone fault (TZ). Within the North American plate, minor strike-slip motion occurs along the Denali fault (DF).
2. Previous Studies

2.1. Southeast Alaska

Hicks and Shofnos [1965] studied uplift in SE Alaska using data from permanent gauges at Skagway, Juneau, Sitka and Ketchikan, along with repeated temporary gauge data at 27 sites distributed throughout northern SE Alaska. A broad, elongate uplift pattern affecting an area of over $2 \times 10^5$ km$^2$ and a peak rate of $>3$ cm yr$^{-1}$ was found centered over Glacier Bay. A tidewater glacier system completely filling Glacier Bay (~1600 km$^2$) which rapidly retreated between 1760 and ~1930. This retreat resulted not only in the vacation of >1 km thick ice from the bay but also in significant drawdown of the icefields that fed the tidewater system, which covered >7000 km$^2$. Hicks and Shofnos [1965] hypothesized that the rapid unloading from this retreat, combined with residual rebound from post-Wisconsin age deglaciation, caused the observed uplift. Since the region is at the beginning of the transition from strike-slip to subduction [e.g., Doser and Lomas, 2000; Fletcher and Freymueller, 1999], others have suggested a tectonic origin or contribution to the uplift. Observations that have been cited include (1) seismic evidence of a possible link between the Fairweather and Denali fault systems trending across the uplift center [Horner, 1983], (2) the breadth of the uplift pattern over southeast Alaska [Hudson et al., 1982], and (3) the lack of changes in gravity measurements that would be consistent with isostatic rebound in the region [Barnes, 1984].

2.2. 1964 Earthquake Near-Field

Sea level observations were used to determine the coseismic offsets in the 1964 Great Alaskan earthquake [Plafker, 1971; Brown et al., 1977], and new tide gauges at Valdez, Seldovia, Nikiski and Anchorage were established in the affected area specifically to study the postseismic vertical motion. Brown et al. [1977] and Savage and Plafker [1991] studied the postseismic deformation in the affected area using annual MSL data. In addition to having a longer set of data available, Savage and Plafker [1991] corrected these data with an average of detrended, concurrent sea level fluctuations observed at three stations in SE Alaska (Sitka, Juneau and Ketchikan). This approach removes oceanographic-related variations in the sea level records, and was found to reduce the scatter in sea level records at stations from Yakutat to Sand Point. At all seven sites in the 1964 near-field, vertical land motion was found to be recovering the coseismic offset, but with a much longer timescale (perhaps ~100 years) than has been found for postseismic deformation in Japan (~2 years) [Kato, 1983]. Cohen and Freymueller [2001] re-examined this data set with 10 years additional data. A significant result was the determination of non-linear trends in vertical crustal motion at Kodiak, Cordova and possibly Valdez.

3. Strategy

The challenge in using sea level data to measure crustal motion is that of comparing one dynamic surface or reference plane with another. The sea surface exhibits extreme variability at high frequencies (e.g., semi-diurnal tidal amplitudes are as much as 10 m in some areas covered by this study), along with small long-term changes (a global average sea level rise of ~1.5 mm yr$^{-1}$ over the last century [Douglas, 1997; Tamisiea et al., 2001]). On the other hand, solid earth vertical motion rarely exceeds 10 mm yr$^{-1}$ and is usually devoid of high frequency variability, with the exception of coseismic offsets, which exceeded 10 m in the 1964 rupture zone [Plafker, 1971]. Vertical crustal motion can have long period variability due to factors such as postseismic relaxation, passive margin subsidence and glacial isostatic adjustment (GIA).

Two basic approaches generally have been employed to counter the high frequency variability of the sea surface in order to obtain a stable reference plane: (1) long-term averaging of sea level data and, (2) various schemes to identify and remove oceanographic “noise.” Although sea level averaged over yearly or longer periods will cancel most of the high-frequency variability, long-term averages cannot resolve rapid datum changes, such as coseismic offsets. Because the signal we are interested in has both high-frequency components and long period variability, we use monthly MSL analysis. For many of the older records, this sampling frequency represents a practical limit, as sea level data averaged over periods less than a month can be difficult to obtain. The primary goal of our analysis is to remove sea level variations due to oceanographic and atmospheric effects, and to do so without smoothing the records with long-term averages.

4. Data

The National Ocean Service (NOS) in Alaska and the Marine Environmental Data Service in Canada collected the sea level data used in this study. These data are typically collected from an instrument whose reference plane stability is verified by annual leveling surveys, thereby insuring a consistent datum between sea level measurements and the land level. With long-term records, this issue is complicated by changes in instrument location and, in the case of the 1964 earthquake, destruction of the instrument platforms by tsunamis. The data providers have corrected most of the records for these issues; in the case of Kodiak we have...
incorporated leveling data and simultaneous water level observations to link together records from tide gauges at two different locations [Cohen and Freymueller, 2001]. We have not attempted to recover the large coseismic datum shifts at Seward and Kodiak associated with the 1964 earthquake, but instead break these records at the time of the earthquake for coseismic offsets, see [Plafker [1971] and Brown et al. [1977]].

The record lengths for each station are summarized in Table 1. Almost all of the stations have data gaps of variable length within their records. In addition to periods of malfunctioning equipment, these data gaps may represent periods of datum instability. In a few instances this instability is a signal we are interested in, such as the immediate postseismic behavior following the 1964 earthquake. In these instances a much more continuous complete record of mid-tide level (MTL) is sometimes available. MTL is the mean of all high and low water levels (the peaks of a tidal series), while MSL is simply the mean of all water level readings. At Anchorage and Kodiak, the MSL records are not available for several years following the 1964 Great Alaskan earthquake. At Cordova, there is a gap of two years in the MSL record, 1968–1969. The MTL records for these stations were used in our analysis.

5. Sea Level Analysis

Our analysis differs from that of previous uplift studies using sea level records in the Gulf of Alaska by (1) using monthly mean sea levels rather than annual means, (2) correcting each station for barometric variations and seasonal variations, and (3) using a much larger set of stations to estimate common-mode oceanographic variations. Figure 2 illustrates our three-step analysis process using the Yakutat and Skagway records as examples. Of primary concern is the precise removal of all signals within these records that are not related to crustal motion.

5.1. Barometric Pressure Corrections

Starting with the raw monthly mean sea levels ($MSL_{i}$, where the subscript refers to the ith month in a multiple year record), the first step is to correct for barometric pressure variations. Monthly mean sea level air pressure values ($P_i$) were found for each tide gauge site from the NCEP reanalysis of climatological data, 1949–2001 (NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado, USA; http://www.cdc.noaa.gov). Records from before 1949 are not corrected for barometric pressure variations. At each station, we calculated the mean barometric pressure ($\bar{P}$) over the 52-year period. Sea level variations due to pressure variations about $\bar{P}$ were then removed, assuming 1.0 cm sea level change per mbar, to find pressure corrected sea levels ($PC_i$).

\[ PC_i = MMSL_i - 1.0 \frac{\text{cm}}{\text{mbar}} \times (P_i - \bar{P}) \]  

Other authors have found it more effective to use a larger rate than $-1.0 \text{ cm mbar}^{-1}$ to remove all of the correlated sea level and air-pressure variations, possibly due to correlation of effects such as wind loading with barometric pressure variations [Trupin and Wahr, 1990; Davis et al., 1999]. However, we found that $-1.0 \text{ cm mbar}^{-1}$ provided the greatest RMS noise reduction about the long-term trends for the stations studied here.

5.2. Removal of Seasonal Signal

All of the stations exhibit seasonal sea level variations to a variable degree. This seasonal signal is due to a combination of several effects, including seasonal variations in air and water temperatures, predominant wind direction, and fresh water influx resulting in local density variations. Particularly strong seasonal signals are evident at Anchorage and Skagway; both of these sites lie at the head of long inlets. In general, of the sites we studied, those closer to the open ocean appeared to have weaker seasonal signals. To correct the long-term records for each of the specific local effects that cause these seasonal signals would require detailed analysis involving wind, temperature, precipitation and runoff data that are either difficult to obtain or do not exist.

Instead, we find an average seasonal signal at each station, and remove this signal from the record [Kato, 1983]. To calculate each station's monthly deviation $D_{i}$ from the overall trend, we run a high pass filter (Chebyschev type II, 10 year cutoff) on each station’s pressure corrected record, producing the equivalent of monthly MSL deviations from a smooth ten-year trend. The average deviation of each month ($\overline{D_m}$; $m = 1$ to 12), over the entire record of this particular month (N years of data for month m) is calculated:

\[ \overline{D_m} = \frac{1}{N} \sum_{j=0}^{N-1} D_m \]  

These average monthly deviations are subtracted from each matching month throughout the record, thus finding “seasonally corrected” sea levels ($SC_{i}$):

\[ SC_{i_j} = PC_{i_j} - \overline{D_m} \]  

for $m = 1$ to 12 and $j = 1$ to N. It should be noted that there is no analogous step for analysis of annual MSL data, although in the long-term records data gaps of one to several years are not available for several years following the 1964 Great Alaskan earthquake. At Cordova, there is a gap of two years in the MSL record, 1968–1969. The MTL records for these stations were used in our analysis.
months are common. Annual means calculated over these gaps will be weighted by the seasonal signal of the remaining points, introducing a bias for any year without a complete set of monthly means.

5.3. Removal of Common Mode Oceanographic Signal

The standard procedure to remove sea level variations caused by large-scale oceanographic variations is as follows. A linear sea level trend is found for each stable, high quality site within the study area, and monthly residuals about these trends are tabulated \( (R_s)_i \), where \( s = \) site and \( i = \) month). These residual series are then combined to form an average residual for each month. This monthly residual average is taken to represent regional sea level variations that are common to all sites \([\text{Kato}, 1983; \text{Davis} \text{ et al.}, 1999; \text{Cohen} \text{ and Freymueller}, 2001]\), and is called the “oceanographic correction” (OC). The stations used to form this average monthly residual are called “control stations.” At each station, for every month \( i \), \( \text{OC}_i \) is subtracted from \( \text{SC}_i \) to find sea level fully corrected by our analysis \( (H_i) \). Each \( H_i \) is calculated with an \( \text{OC}_i \) found over a control station subset \( s = 1 \) to \( k \) as follows:

\[
H_i = \text{SC}_i - \text{OC}_i, \quad \text{where} \quad \text{OC}_i = \frac{\sum_{s=1}^{k} R_s}{k} \]

The specific control stations and their total number \( (k) \) may vary from month to month depending on data availability. We restrict the computation of \( H_i \) to only those months when three or more control stations’ residuals are included in \( \text{OC}_i \). This requirement limits our complete analysis to 1937 and later.

The above procedure makes a limiting assumption that the MSL trends are linear at all the sites used to compute the OC. Because of known instabilities and non-linear trends in many of the subduction zone sites, previous sea level studies in this region have restricted the computation of the OC to include only residuals \( [R_s] \) from three sites in SE Alaska (Juneau, Sitka and Ketchikan) which have linear sea level trends \([\text{Savage and Plafker, 1991; Cohen and Frey-}\]

Figure 2. The primary steps in our tidal record analysis, with Skagway and Yakutat as examples. Skagway has a strong and difficult to correct for seasonal signal. Our analysis is more effective at Yakutat, revealing significant information of the uplift history.
mueller, 2001]. This approach is less than ideal, as the computation of the OC with only a few stations occasionally introduces strong local sea level variations that might be evident at one station only. An average of $[R_s]_p$, computed over many stations will damp out individual station variations, and better resolve those variations common to all sites.

[21] Consequently, to strengthen the OC, we use an analysis developed by Kato [1983] to incorporate residuals from non-linear subduction zone stations. Instead of finding residuals about a linear trend, a low-pass filter is used to approximate the long-term trend. The residuals $[R_s]_p$ are then calculated about this trend. Furthermore, we have included in our analysis two additional stations (Queen Charlotte City and Prince Rupert). Together, these additions allow the number of control stations to be increased from three as used by Savage and Plafker [1991] and Cohen and Freymueller [2001] to 11 (present study).

[22] For stations in the strike-slip regime, a Chebyshev type II filter with a 50 year cutoff and 20 dB down in the stop band was used to approximate the long-term trend. The long period of the cutoff frequency is appropriate for these predominantly linear stations. This cutoff frequency allows for long period oceanographic effects to be included in the OC, such as the lunar nodal tide which has an 18.6 year frequency and an amplitude up to $\sim 50$ mm [Trupin and Wahr, 1990]. For stations within or near the subduction zone (all stations north of Skagway), such a long period cutoff is not allowable as the solid earth uplift rates may vary on a faster timescale. Here, we used the same filter to estimate trends as in the strike-slip regime, but with a cutoff of 10 years [Kato, 1983].

5.4. Coseismic Offsets

[23] Sudden jumps or datum shifts, such as those during great earthquakes, are not removed by low-pass filtering. The long-term trend approximated by low-pass filtering will gradually span these sudden shifts, and the non-physical residuals from this poorly approximated trend will then contaminate the OC. To correct this situation, we break certain records into separate pieces. The 1964 Great Alaskan earthquake (Mw 9.2 [Kanamori, 1977]) caused uplift of roughly 1.0 m at Seward and 1.2 m at Kodiak [Plafker, 1971], and the records at these stations are treated separately before and after the earthquake. In the strike-slip regime, there have been three great earthquakes this century: 1958 Fairweather fault, Mw = 8.2 [Kanamori, 1977], 1949 Queen Charlotte fault, Mw = 8.1 [Kanamori, 1977], and 1972 Sitka, Mw = 7.6 [Schell and Ruff, 1989]. In general, however, strike-slip events result in small vertical displacements in the far-field, and in the analysis presented here all records have been treated as continuous across these events.

5.5. Final Selection of Control Stations

[24] With the earthquake-correlated datum shifts identified, and the records separated at these times, all stations from the entire study area that have long-term and relatively clean records can contribute their residuals to the OC calculation. We did not include several records in the OC for the following reasons. The Skagway record has a higher than average noise level, perhaps due to a particularly strong seasonal signal. Yakutat has a complicated non-linear trend, and may have several minor offsets (see section 7.3).

6. Results

[25] The sea level analysis outlined above was applied to the 15 stations within our study area. We have calculated relative sea level rates for these sites and determined some to be non-linear; the results of our linear/non-linear rates analysis are presented in Figure 3. These records have been adjusted with the approximate local effect of global sea level rise (1.0 mm/yr), as discussed below, resulting in the vertical crustal motion records presented in Figures 4 and 5.

6.1. Measurement Error

[26] An exact value for the measurement error associated with monthly mean sea levels (MMSL) is difficult to determine. This is not so much due to the difficulty in accurately measuring individual water levels, rather it is the accuracy of the mean reference plane obtained from the highly variable sea surface that is of concern. This accuracy can only be found directly from the data and not from independent error estimation. We employed a high-pass filter with a 5 year cutoff to determine the variance of each of the records, both of the raw data and the fully corrected sea level $H_s$. The reduction of variance ($\sigma^2$), averaged over all the control stations, is a factor of 18.5 between the raw and fully corrected sea level records. In the analysis that follows, we will use a constant MMSL measurement error estimated from the standard deviation about the 5 year high-pass filtered records from each of the control stations.

[27] The average standard deviation of corrected sea level records at the 11 control stations is 20.9 mm, with two stations having significantly smaller values: Sitka 12.5 mm and Cordova 14.1 mm. These two sites may represent the best case for our analysis, with both sites being close to the open ocean and in the center of the distribution of control stations. We might expect Yakutat to have a similarly low $\sigma$, but it is 19.8 mm, possibly because it may contain several small offsets that are admitted by the 5 year high-pass filter (see section 7.3).

6.2. Constant and Time-Dependent Rate Analysis

[28] To determine relative sea level rates at each station, each MMSL record was fit with linear and time-dependent regressions. To identify statistically valid time-dependence, we assessed the improvement in fit of a time-dependent regression over that of a linear regression. The statistical significance of each regression was evaluated using the $F$ test [Zhao et al., 1995] and reduced $\chi^2$. We tested logarithmic and exponential forms, motivated by glacial isostatic models, as well as by postseismic afterslip and viscoelastic models. At none of the stations was there significant improvement in fit with these forms over that
<table>
<thead>
<tr>
<th>Tide Gauge</th>
<th>Linear Stations</th>
<th>Non-Linear Stations</th>
<th>Reduced $\chi^2$ (of preferred model)</th>
<th>Linear vs. Quadratic model comparison</th>
<th>Comments</th>
</tr>
</thead>
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<tr>
<td></td>
<td>RSL = A*(t-t_0) + B*(t-t_0)^2</td>
<td></td>
<td>F Ratio</td>
<td>Improvement in reduced $\chi^2$</td>
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<tr>
<td>Queen Charlotte City</td>
<td>0.0 ± 0.1</td>
<td>A (mm/yr) B (mm/yr$^2$) t_0 (yr)</td>
<td>2.0</td>
<td>9.1</td>
<td>1.9%</td>
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<tr>
<td>Prince Rupert</td>
<td>1.72 ± 0.06</td>
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<td>1.3</td>
<td>0.7</td>
<td>0.0%</td>
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<tr>
<td>Ketckikan</td>
<td>0.02 ± 0.04</td>
<td></td>
<td>1.4</td>
<td>28.8</td>
<td>3.7%</td>
</tr>
<tr>
<td>Sitka</td>
<td>-2.03 ± 0.03</td>
<td></td>
<td>0.5</td>
<td>0.1</td>
<td>-0.1%</td>
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<tr>
<td>Juneau</td>
<td>-12.63 ± 0.06</td>
<td></td>
<td>1.9</td>
<td>6.1</td>
<td>0.7%</td>
</tr>
<tr>
<td>Skagway</td>
<td>-16.3 ± 0.2</td>
<td></td>
<td>10.7</td>
<td>42.7</td>
<td>6.2%</td>
</tr>
<tr>
<td>Yakutat</td>
<td>-5.76 ± 0.06</td>
<td>-0.061 ± 0.004</td>
<td>1970.5</td>
<td>1.8</td>
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<tr>
<td>Cordova</td>
<td>4.94 ± 0.15</td>
<td>-0.23 ± 0.02</td>
<td>1986.75</td>
<td>0.8</td>
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<td>Valdez</td>
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<td>-0.39 ± 0.02</td>
<td>1986.75</td>
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<td>Seward</td>
<td>-1.4 ± 0.3</td>
<td>-0.26 ± 0.03</td>
<td>1986.75</td>
<td>1.4</td>
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<tr>
<td>Seldovia</td>
<td>-8.6 ± 0.2</td>
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<td>1.4</td>
<td>13.0</td>
<td>3.5%</td>
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<td>Nikiski</td>
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<td>0.43 ± 0.04</td>
<td>1986.75</td>
<td>2.3</td>
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<tr>
<td>Anchorage</td>
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<td>-0.24 ± 0.04</td>
<td>1986.75</td>
<td>2.7</td>
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<td>Kodiak</td>
<td>-15.1 ± 0.2</td>
<td>0.33 ± 0.02</td>
<td>1984.5</td>
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<tr>
<td>Sand Point</td>
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<td></td>
<td>1.7</td>
<td>5.2</td>
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<tr>
<td>Seward (Pre-64)</td>
<td>-1.0 ± 0.3</td>
<td></td>
<td>2.3</td>
<td>18.1</td>
<td>7.9%</td>
</tr>
<tr>
<td>Kodiak (Pre-64)</td>
<td>-1.1 ± 0.5</td>
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<td>1.4</td>
<td>18.9</td>
<td>10.1%</td>
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<tr>
<td><strong>Tri-Linear and Cubic Analysis (Yakutat Only)</strong></td>
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<tr>
<td></td>
<td>A (mm/yr) B (mm/yr$^2$) C (mm/yr$^3$)</td>
<td>Reduced $\chi^2$</td>
<td>Linear vs. Cubic model comparison</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Yakutat</td>
<td>-4.02 ± 0.12</td>
<td>-0.06 ± 0.003</td>
<td>-0.003 ± 0.0002</td>
<td>1.4</td>
<td>259</td>
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<tr>
<td><strong>Linear Rate 1940 to 1953</strong></td>
<td>Reduced $\chi^2$</td>
<td>Linear Rate 1953 to 1979</td>
<td>Reduced $\chi^2$</td>
<td>Linear Rate 1979 to 2001</td>
<td>Reduced $\chi^2$</td>
</tr>
<tr>
<td>Yakutat</td>
<td>-7.4 ± 0.6</td>
<td>1.5</td>
<td>-3.1 ± 0.2</td>
<td>1.4</td>
<td>-10.4 ± 0.2</td>
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<tr>
<td><strong>PSMSL Record Linear Rate 1979 to 2001 with 35 mm step at Jan 1993 removed</strong></td>
<td>Reduced $\chi^2$</td>
<td></td>
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</tr>
<tr>
<td>Yakutat</td>
<td>-12.7 ± 0.2</td>
<td>1.0</td>
<td></td>
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**Figure 3.** Permanent tide gauge analysis.
of a quadratic regression, and therefore quadratic regressions were used for our constant versus time-dependent rate evaluations.

To compare one model with another, we compute an $F$ ratio [Zhao et al., 1995; Cohen and Freymueller, 2001]. Eleven of the records passed the $F$ test at the 0.1% level (Figure 3) indicating that the quadratic model provides a better fit than a linear one. This conclusion includes a few sites that would seem to be better represented by a linear model, such as Queen Charlotte City. However, another test, calculated using $\ln(\chi^2_{\text{linear}}/\chi^2_{\text{quadratic}})$, indicates that a few of the records passing the $F$ test are perhaps not better fit by a quadratic regression. Based on the combined results of the $F$ test and improvement in reduced $\chi^2$, we find that a quadratic model is an improvement over a linear one for Yakutat, Cordova, Valdez, Seward, Nikiski, Anchorage and Kodiak (Figure 3). Perhaps the most noticeable result of our
time-dependent analysis is that our model comparisons indicate two distinct groups. Stations that pass our criteria for non-linear trends tend to pass strongly, while the remaining stations either weakly pass one test only or fail both. The results of our preferred model for each site are presented in Figure 3. The regressions and 95% confidence intervals for these models are plotted as dashed lines in Figures 4 and 5.

The criteria used here are rather stringent; we require that the quadratic models pass the $F$ test at the 0.1% level ($F$ ratio $>$10.8) [Beyer, 1991] and show an improvement in $\chi^2$ of $>$10%. Our motivation for such stringent criteria is to compensate for the possibility that temporal correlations may be introduced by our analysis. We note that in our statistical tests we invoke the assumption that each corrected tidal series is composed of independent variables, and that our models predict a functional relationship between measured and model-dependent variables. This assumption of

Figure 4. Uplift versus time, reduced from monthly mean sea level records at tide gauge stations in the southern half of our study area. Upward land motion is shown as positive. Solid lines show the regressions listed in Figure 3, with 1 mm/yr geoid rise added (see text), and the dashed lines show the 95% confidence intervals.

Figure 5. Uplift versus time, reduced from monthly mean sea level records at tide gauge stations in the northern half of our study area. Upward land motion is shown as positive. Solid lines show the regressions listed in Figure 3, with 1 mm/yr geoid rise added (see text), and the dashed lines show the 95% confidence intervals.
measured independent variables is not strictly valid, as we have already applied a model for removing the seasonal, oceanographic and atmospheric signals.

In addition to comparisons between linear and quadratic models for all the sites, we considered further models at Yakutat and Kodiak. The Yakutat record has an obvious double curvature (Figure 4), and a cubic model was found to be a significant improvement over the already justified quadratic fit (Figure 3). We also tested a linear regression separated at 1953 and 1979, with three independent segments. This model further improved the fit; although the number of model parameters is now six (three slopes and intercepts), the increase easily passed the F test (Figure 3). The final rate presented for Yakutat in Table 2 is the slope of the last of these three linear segments. Within this segment, we have removed an abrupt shift in January 1993 that may be associated with an instrument change that occurred then (see Yakutat discussion, below).

As discussed earlier and also by Cohen and Freymueller [2001], the post-1964 Kodiak record is composed of data from two sites, St. Paul Harbor (1967–1982) and Women’s Bay (1985 to present). We tested a model composed of two independent linear regressions separated at the time of the gauge relocation. We found that a quadratic model over the entire period resulted in a significantly better fit than either a single- or bi-linear regression, and therefore agree with Cohen and Freymueller [2001] that the time dependency of the complete uplift record at Kodiak post-1964 is unlikely due to differential uplift at the two sites.

### 6.3. Relative Sea Level Trends and Uplift Rates

To use relative sea level records as a proxy for vertical crustal motion, one must correct for concurrent rate of change in the geoid (absolute sea level). Globally, sea level is rising an average of ~1.5 mm/yr [Douglas, 1997; Tamisiea et al., 2001], but locally this value will differ near changes in surface mass, e.g. melting glaciers and ice sheets [Woodward, 1888; Farrell and Clark, 1976; Mitrovica et al., 2001; Tamisiea et al., 2001]. Glaciers and icefields in Alaska and neighboring Canada have lost an average of 42 Gt yr\(^{-1}\) from the mid-1950s to the mid-1990s [Arendt et al., 2002]. To find the associated and site-specific geoid changes at each of our tide stations would require sophisticated calculations incorporating detailed geometry of the surface mass changes. Instead, we use the results of a study of non-eustatic sea level redistributions by Tamisiea et al. [2001] to estimate the effect of global sea level rise on our study area. In addition to models of ice mass changes in Greenland and Antarctica, Tamisiea et al. [2001] considered ice mass fluctuations in a suite of mountain glaciers [Meier, 1984] that includes an approximation of Alaskan ice mass changes that agrees reasonably well with recent measurements [Arendt et al., 2002]. The rate of global sea level rise (\(g\)) predicted for our study area is \(\sim 1.0\) mm yr\(^{-1}\) [Tamisiea et al., 2001]. Although the formal error estimate for this value is \(< 1\) mm yr\(^{-1}\), we conservatively assign an error of \pm 1 mm yr\(^{-1}\) due to the uncertainty involved with using this value to convert relative sea level trends to uplift records.

In addition to the effect of surface mass changes, motion of the Earth’s mantle can also cause geoid variations. Such an effect may be present in our study area from viscoelastic postseismic relaxation [Wahr and Wyss, 1980; Pollitz et al., 2001] and glacial isostatic adjustment. Measuring or modeling the site-specific magnitude of the related geoid change is outside the scope of our present study, and we do not adjust the uplift records for either.

### 6.4. Global Glacial Isostatic Adjustment

Global glacial isostatic adjustment (GIA) models are poorly resolved for our study area due to limited knowledge of last glacial maximum ice-sheet history here [e.g., Tushingham and Peltier, 1991]. To the south of our study area in the northern Cascadia subduction zone, James et al. [2000] constructed a regional post-glacial rebound model with a detailed Cordilleran ice sheet history. This model predicts present-day uplift rates less than 0.1 mm/yr, lower than ICE-3G predictions by an order of magnitude [Tushingham and Peltier, 1991]. Owing to this uncertainty, we do not separate out any estimates for GIA from the uplift records. Moreover, ICE-3G uplift predictions throughout our study area are \(< 1\) mm yr\(^{-1}\), a small fraction of the uplift rates at most of the stations here.

### 6.5. Present-Day Uplift Rate Predictions

Using the results of the constant and time-dependent analysis above, we calculated the relative sea level rate for each site predicted by the preferred model for the year 2000. We estimate the uplift rate for each site by subtracting the estimated rate of change in the geoid (Table 2). The errors listed are the quadrature sum of the MSL trend error (site dependent) and the geoid rate error (\pm 1 mm yr\(^{-1}\)). For all sites but Seward, Nikiski, and Anchorage the MSL trend errors are insignificant compared to the geoid rate error.

### 7. Discussion

#### 7.1. Southern Stations

The southernmost stations in the strike-slip regime (Queen Charlotte City, Prince Rupert, Ketchikan and Sitka) all show linear, slow uplift trends (Figure 4). Prince Rupert and Ketchikan both have uplift rates close to zero. Queen Charlotte City is within 60–70 km of the Pacific Plate transform boundary at this latitude, and the small observed uplift might be associated with this close proximity. With the relatively small uplift trends at these southern three stations, glacial isostatic adjustment following the last glacial maximum may be significant. Sitka exhibits a faster
uplift rate that represents the southern periphery of the SE Alaska uplift area.

7.2. Southeast Alaska Uplift Region

[38] The next three stations to the north (Juneau, Skagway and Yakutat) all have extremely rapid uplift rates. There have been several suggestions for the cause of rapid uplift in this region: (1) isostatic rebound following the rapid deglaciation of Glacier Bay [Hicks and Shofnos, 1965; Motyka, 2003], (2) both this and ongoing regional deglaciation [Clark, 1977], (3) regional tectonic stress [Horner, 1983; Barnes, 1984] or (4) some combination of these [Hudson et al., 1982].

[39] Recent measurements of relative plate motion have found predominantly strike slip motion along the Pacific-North American plate boundary to the south of Yakutat, implying that any compressive motion must occur offshore on the Transition Zone fault [Fletcher and Freymueller, 1999; Larsen et al., 2001]. These observations suggest that tectonic deformation is likely a minor component of the total uplift rates, and we will concentrate here on testing isostatic rebound models. Our approach is to create a model of surface load changes from direct and indirect observations of ice mass changes, and calculate the resultant uplift. Comparisons between these rate predictions and observed rates are then used to evaluate the models. We use the most recent rate at Yakutat (1979 to present, with an offset at 1993 removed; see section 7.3) under the assumption that earlier in the record the uplift there was modulated by tectonics. Our goal with the modeling efforts presented here is not to precisely constrain material properties of the crust and mantle, but rather to demonstrate that realistic models of glacial unloading, coupled with an established Earth model, can indeed produce uplift of the magnitude observed in southeast Alaska.

[40] The Earth model used for these calculations is described in detail by Ivins and James [1999], and is a flat-Earth, gravitating, incompressible two-layer model with an elastic lithosphere and a viscoelastic mantle half-space. The geometry and material parameters of this model are shown in Figure 6. This simple model is sufficient to evaluate the effects of regional ice load variations over <20° of Earth surface [Ivins and James, 1999]. For simplicity, we limited our calculations to an Earth model with 40 km thick crust [Bechtel et al., 1990]. We will consider load variations related to (1) ongoing ice thickness changes in Alaska and neighboring Canada [Arendt et al., 2002], (2) a ~1000 year regional ice thickness history based on dendrochronology and geomorphology studies of glacial advance and retreat [Wiles et al., 1999], and (3) the most recent cycle of advance and retreat of the tidewater glaciers in Glacier Bay [Goodwin, 1988]. We do not consider ice mass changes older than 1000 years, nor any ocean loading due to sea level changes.

[41] The present-day ice mass loss of Alaskan and neighboring Canadian glaciers is estimated at 96 ± 35 Gt/yr from direct measurements of ice thickness changes [Arendt et al., 2002]. The associated ice thickness changes vary strongly with altitude, and the greatest ice loss occurs at the lowest elevations. To create a load model based on these observations, we begin with a 3-arc second (90 m resolution) digital elevation model of the ice covered area, and find the ice thickness changes at each node using an average thickness change versus elevation profile [Arendt et al., 2002, Figure 2b]. Because this is a regional average profile, we do not account for the often-large differences in mass changes specific to tidewater glaciers. To facilitate computation of the Earth’s isostatic response to this load model, we average the load changes over a 20 km × 20 km grid. The grid dimensions are chosen to be about half of the effective elastic crustal thickness here [Bechtel et al., 1990].

[42] We first calculate the elastic isostatic response to this load model (η ≈ ∞). We find that the elastic uplift rates are only ~10–30% of the observed rates at Sitka, Juneau, Skagway and Yakutat (Table 3). The uplift error estimates in Table 3 are calculated from models using the upper (131 Gt/yr) and lower (61 Gt/yr) limits of the current ice mass loss rate [Arendt et al., 2002].

[43] To perform a viscoelastic calculation, a history of ice load changes is required. The Little Ice Age (LIA) glaciation was the largest Holocene expansion in Alaska, and occurred in three phases between ~1200 and 1900 AD that were synchronous across much of Alaska [Wiles et al., 1999; Calkin et al., 2001]. Our interpretation of the associated changes in ice mass through the last 1000 years is shown in Figure 7. The results of Arendt et al. [2002] are used from 1955 to the present. This load model includes the ongoing rate of load change from 1995 to the present day, and the elastic uplift results of the previous calculation are retained in the following predictions. The rate of load change for 1900 to 1955 is assumed to be the same as the 1955–1995 rate [Arendt et al., 2002]. Prior to 1900, the maxima and minima are fractional estimates relative to the 1900 peak. Again, this is a regional average and it does not account for present or past mass changes of tidewater glaciers. For sensitivity analysis, we used the error limits on the volume change rates [Arendt et al., 2002] to extrapolate minimum and maximum load history estimates (Figure 7).

[44] Little or no uplift was predicted with this load model using mantle viscosities similar to those found in Fennoscandia (η ≈ 10^{21} Pa s) [e.g., Milne et al., 2001]. Earth models with viscosities >10^{20} Pa s primarily respond at the present day, if at all, to the growth phase of the load history.
Table 3. Glacial Isostatic Rebound Model Uplift Rate Predictions

<table>
<thead>
<tr>
<th>Station</th>
<th>Elastic Uplift Only mm/yr</th>
<th>Regional LIA Viscoelastic $\eta = 3.5 \times 10^{19}$ Pa s mm/yr</th>
<th>Regional Plus Glacier Bay Viscoelastic $\eta = 3.5 \times 10^{19}$ Pa s mm/yr</th>
<th>Observed Uplift Rate, mm/yr</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sitka</td>
<td>0.9 ± 0.2</td>
<td>4.3 ± 1.0</td>
<td>6.4 ± 1.1</td>
<td>3.0 ± 1.0</td>
</tr>
<tr>
<td>Juneau</td>
<td>1.6 ± 0.5</td>
<td>7.9 ± 1.8</td>
<td>12.6 ± 2.1</td>
<td>13.6 ± 1.0</td>
</tr>
<tr>
<td>Skagway</td>
<td>1.9 ± 0.6</td>
<td>9.4 ± 2.1</td>
<td>17.9 ± 3.0</td>
<td>17.1 ± 1.0</td>
</tr>
<tr>
<td>Yakutat</td>
<td>2.7 ± 0.7</td>
<td>14.5 ± 3.1</td>
<td>16.8 ± 3.2</td>
<td>13.7 ± 1.0</td>
</tr>
</tbody>
</table>

used here, and would not visously respond to the retreat phase until the future. In contrast to the high viscosity model results, significant present-day uplift rates are obtained with low mantle viscosities ($\eta \approx 2 \times 10^{19}$–$5 \times 10^{19}$ Pa s). Results for our best-fit model with $\eta = 3.5 \times 10^{19}$ Pa s are presented in Table 3, along with a range of uplift rates from models incorporating the maximum and minimum load histories of Figure 7. Although these predictions are close to the observed rates at Sitka and Yakutat, the disagreement with the observed rates is much larger at Skagway and Juneau. This suggests a deficiency in the spatial distribution of our load model, so we next consider the rapid retreat of Glacier Bay ca. 1750–1930 AD.

To create a load model of the most recent cycle of advance and retreat in Glacier Bay, we used an estimate of the above sea level ice mass at the ~1750 ice extent. Trimlines and lateral moraines provide indications of post-1750 ice thickness changes within Glacier Bay [Field, 1947; Clague and Evans, 1993]. We used present-day tidewater glacier analogs to estimate ice elevation profiles within Glacier Bay using these thickness changes, extending ice coverage out to the 1750 terminus position. Comparing these ice profiles with a digital elevation model, we find that Glacier Bay lost a minimum of 2600 km$^2$ of ice from 1750 to ~1950 AD, at an average unloading rate of $\sim1.2 \times 10^{13}$ kg/yr. This load change is partitioned into the five disk loads shown in Figure 8, and the load history of these disks, based on Goodwin [1988], is shown in the inset.

When the Glacier Bay load model is combined with the regional LIA load model, the uplift rates predicted with a low viscosity Earth model closely match the observations at Sitka and Yakutat, but overshoot the observations at Juneau and Skagway, by ~3 mm/yr (Table 3, limits are from varying the Glacier Bay load model by ±25%, combined with range of rates from the regional LIA load model). Our results indicate that the significance of the Glacier Bay load relative to the LIA regional load is ~37% at Juneau and ~47% at Skagway, while perhaps it is only a minor contribution at Sitka and Yakutat. The regional pattern of uplift we predict (Figure 8) is in reasonable agreement with that presented by Hicks and Shofnos [1965] from temporary tide gauge measurements.

With the long history of active tectonics and nearby volcanism of the region, a low upper mantle viscosity is conceivable. Southeast Alaska has been subjected to dextral strike-slip motion along the North American-Pacific plate boundary, in a configuration similar to the present day situation, for ~25 Myr [Plafker, 1987]. Quaternary volcanism is found at Sitka (Mt. Edgecombe) and ~200 km to the north of Yakutat (Wrangell Mts.). Elsewhere along the North American-Pacific plate boundary, studies of postseismic deformation in southern California [Pollitz et al., 2001] and glacial isostasy in southern British Columbia [Clague and James, 2002] have indicated upper mantle viscosities on the order of $1 \times 10^{19}$ Pa s. The relatively brief load history considered here is appropriate for Earth models with such low viscosities, as the response to earlier load changes will have largely been completed by the present. With these considerations, we regard the modeling results as realistic and plausible, and consider the rapid uplift observed in much of southeast Alaska to be attributable to viscoelastic post-glacial rebound following the LIA and the tidewater retreat of Glacier Bay.

7.3. Yakutat

[48] In addition to the high uplift rate, there is a strong degree of non-linearity in the record at Yakutat. Here, in the transition zone between strike-slip and subduction, there can be no question of tectonic influences. Yakutat lies on the Yakutat Block, an allochthonous terrane actively colliding with North America [Plafker et al., 1994]. The magnitude of horizontal velocity at Yakutat is almost equal to the full Pacific plate velocity relative to North America [Fletcher and Freymueller, 1999]. About 110 km northwest of Yakutat the plate boundary changes from strike-slip motion to convergence in a complex transition of steeply-dipping stacked thrust-faults [Bruns and Schwab, 1983] with a convergence rate twice that of the Himalaya (J. Freymueller, unpublished GPS data).

[49] The Yakutat record shows a complex non-linear uplift pattern, with a period of slower uplift from roughly

Figure 7. Regional Little Ice Age load model history. The vertical line at 1955 indicates when direct observations of regional glacial thickness changes begin [Arendt et al., 2002]; prior to this date the load history is estimated using geomorphic records of glacial termini advance and retreat as a proxy for mass changes (see text). Minimum and maximum load histories, based on the error limits of the post-1955 direct measurements, are shown as dotted lines.
1953 to 1979 (Figure 4). Before 1953 the uplift rate at Yakutat was 7.8 ± 0.5 mm yr⁻¹, nearly twice the linear uplift rate of 4.2 ± 0.2 mm yr⁻¹ from 1953 to 1979. The variance about a linear trend between 1953 and 1979 is larger than elsewhere in the record, with two possible shifts at 1958 and 1962. The former may be coseismic uplift on the order of 2–5 cm at the time of the 1958 Fairweather fault Ms = 7.9 earthquake, but is not distinctly above the variance of the record here. The uplift record does not show a significant coseismic signal at the time of the 1964 Great Alaskan earthquake, which initiated 200 km to the northwest in Prince William Sound (PWS).

After the 1979 St. Elias earthquake (Mw = 7.4) [Lahr and Plafker, 1980], the apparent linear uplift rate at Yakutat is 11.0 ± 0.3 mm/yr. The change in uplift rate in 1979 is the most prominent feature in the Yakutat record, and some association with the 1979 St Elias earthquake seems likely. However, a postseismic response of greater than 20 years for an earthquake of this magnitude is markedly different than observations elsewhere. Subduction zone earthquakes of similar magnitude in Japan have postseismic responses lasting ~2 years [Kato, 1983], although in the case of Yakutat the tide gauge resides on the downgoing plate rather than on the overriding plate as in Japan. An alternate explanation for the change in rate at 1979 may be related to the rapid succession of large regional earthquakes (1958, 1964 and 1979). Rather than the post-1979 rate representing lingering effects of that event, it may have returned to a “normal” rate after a period of anomalous tectonic strain between ~1953 and 1979.

In the time period after 1979, the most distinct deviations are a 4–5 year period with decreased uplift rate ending in early 1988, followed immediately by ~1 year of faster than normal uplift. Also there is an approximately 3.5 cm shift downward in January 1994. The former is coincident with the three main events of the 1987–1988 Gulf of Alaska earthquake sequence (11/17/87 Mw = 7.2, 11/30/87 Mw = 7.8, 3/6/88 Mw = 7.7) [Pegler and Das, 1996]. This sequence produced a 13 ± 3 cm downward coseismic offset at Cape Yakataga [Sauber et al., 1993]. The 1994 shift is uncorrelated with any nearby seismicity.

The 1994 shift occurs over one measurement interval only, and the record before and after the shift appears linear. These observations suggest the possibility of a reference level problem rather than crustal motion. We analyzed Yakutat records available from the Permanent Service for Mean Sea Level (PSMSL; http://www.nbi.ac.uk/psmsl) in addition to sea level records directly from NOS. Once fully corrected by our analysis, we found that the Yakutat record...
from PSMSL also had a shift of the same magnitude, but this shift occurs one year earlier in January 1993. This date marks the change of instruments at Yakutat from a float/well system to a NOS next generation water level measuring system. Although the exact source of the differences between the NOS and PSMSL Yakutat records is unclear, we suspect that this 3.5 cm shift is related to the instrument change. With the 3.5 cm shift removed from the uplift record, we find a linear uplift rate between 1979 and 2001 of 13.7 ± 0.2 mm yr⁻¹ (reduced χ² = 1.0). If such treatment of the record is correct, the change in uplift rate in 1979 is even more pronounced.

7.4. 1964 Earthquake Near-Field

[53] The 1964 Great Alaskan Earthquake ruptured two distinct asperities, located beneath Prince William Sound (PWS) and Kodiak Island [Christensen and Beck, 1994; Holdahl and Sauber, 1994]. The tide gauge stations in this region (Figure 1) are distributed along the northeastern edge of the PWS asperity (in order of increasing distance from the megathrust: Cordova, Valdez and Seward), to the west of the PWS asperity (Anchorage), in the region between the PWS and Kodiak asperities (Nikiski and Seldovia), and within the Kodiak asperity (Kodiak). Sand Point, the furthest west station included in our analysis, lies outside of the 1964 near-field and exhibits nearly zero uplift over the entire record (Figures 3 and 5). Geographic correlation of uplift rates to the PWS asperity is marked at stations within the 1964 near-field. Very rapid uplift rates are found outside of PWS, while uplift rates are much slower within PWS (Figure 5). Anchorage is an exception; although west of the PWS asperity, the average uplift rate found over its entire record is relatively small.

[54] At Kodiak, we present an uplift record based on the MTL record that includes the period immediately following the 1964 earthquake (Figure 5). In the 3.5 year period from mid-1964 to 1968, we find extremely rapid uplift totaling 47 ± 8 cm (130 ± 20 mm yr⁻¹). After 1968 the rate decreased, and we find that the uplift trend since this time is non-linear, in agreement with Cohen and Freymueller [2001]. For the quadratic analysis presented in Figure 3, we included data from 1968 onward, on the basis that prior to this a different mechanism of postseismic relaxation was taking place.

[55] In the early record at Anchorage, between 1964 and mid-1972, there is an oscillation with a period of about 5 years. Other authors have speculated on this feature and suggested that it is possibly due to fault creep events propagating slowly along the plate interface [Brown et al., 1977; Savage and Pfafker, 1991]. Cohen and Freymueller [2001] noted that this feature might have been enhanced in previous analyses through biased annual means calculated with less than complete data sets. Our monthly-MTL based analysis avoids this possible bias. NOS conducted regular leveling surveys between the instrument and the local benchmark network and found no evidence of platform instabilities that could create such a signal (S. Lyles, personal communication). Furthermore, the uplift records show similar features at Seward and Seldovia over the same time period (Figure 5), suggesting that the phenomena that caused this oscillatory signal affected a large portion of the 1964 earthquake near-field. The magnitude of the oscillation at Anchorage is 34 ± 5 cm over the first 2.4 years of the record here, corresponding to an uplift rate of 140 ± 20 mm yr⁻¹ over this period. This rate is very similar to the immediate postseismic rate found at Kodiak. Cohen [1998] analyzed leveling surveys along Turnagain Arm between Anchorage and Portage, conducted during the first year of tide gauge operation at Anchorage. He found a maximum uplift, relative to the Anchorage tide gauge benchmark, of ~10 cm. When combined with our results, we find a maximum absolute uplift of 24 cm between May 1964 and May 1965 along the Turnagain Arm leveling profile.

7.5. De-Trended Post-1964 Uplift Records

[56] To illuminate the postseismic responses at Anchorage, Seward, Seldovia, Valdez and Cordova, we have removed the linear uplift trends from these records, and plotted the residual signals in Figure 9. A simultaneous change in the uplift trends at Seldovia, Anchorage and Seward is apparent at 1972.5 (Figure 9). In the period before this change, Seward exhibits an oscillatory behavior similar to Anchorage, but with approximately one half to one third the amplitude. Seldovia also appears to have an oscillatory signal during this period, although at Seldovia the first peak of the oscillation is smaller than the second. The coincident peak at 1972.5 is strongly evident at Seldovia, Seward and Anchorage. To a much lesser extent it may also occur at Cordova, but only slightly above the noise level.

[57] After the 1972.5 peak, four stations (Anchorage, Seward, Valdez and Cordova) exhibit remarkably similar behavior—that of a smooth quadratic trend in uplift rates. The analysis presented in Figure 3 for these stations (as well as for Seldovia) is limited to data from 1972.5 onward; prior to this time we believe the region was undergoing a different response. Kodiak and Nikiski are also in close agreement with each other, while Seldovia is the only station affected by the 1964 earthquake that does not have significant non-linear behavior after 1972.5 (Figure 3).

[58] This pattern in the data suggests that there were two and perhaps three phases in the postseismic deformation following the 1964 earthquake. The initial phase lasted ~3 years, and was characterized by rapid uplift (~130 mm/yr⁻¹) at Kodiak and Anchorage, and even more rapid uplift (~240 mm/year) along Turnagain Arm. A possible second phase was absent at Kodiak but clearly seen around Prince William Sound for another ~5 years, and was characterized by oscillatory vertical motion at sites around the Prince William Sound asperity. The final phase consists of slowly changing uplift rates over the entire area.

[59] Zweck et al. [2002] compared models of postseismic deformation averaged over the ~35 years since the earthquake to models averaged over the last 7 years. The major difference between the two time periods is that the 35 year average model includes rapid afterslip immediately downdip of the 1964 rupture zone that is absent over the last few years. The initial phase of postseismic deformation is plausibly caused by afterslip immediately downdip of the coseismic rupture. This interpretation suggests that the afterslip had decayed within 3–4 years, a similar timescale to that observed for the total postseismic deformation after many other subduction zone earthquakes. The initial phase was followed by a few years of oscillatory motion around Prince William Sound, which may be caused by propagat-
North America-Pacific plate boundary. Our analysis is based on monthly mean sea levels corrected for barometric pressure variations; this differs from previous studies in this region, which have been based on annual mean sea levels. In addition to the increased temporal resolution, this approach allows for improved corrections of local seasonal variations in sea level, and avoids biasing that can occur when annual sea levels are calculated over the data gaps common to the long-term records of this region. This analysis allows us to use a larger number of stations in the determination of common mode oceanographic effects. The increased accuracy of the reduced uplift records enables the statistical determination of uplift rates and trends along this plate boundary.

Figure 9. De-trended uplift records from stations near the Prince William Sound (PWS) asperity of the 1964 Great Alaskan earthquake. Solid lines show the regressions listed in Figure 3, and the dashed lines show the 95% confidence intervals (also de-trended).

8. Conclusions

We have presented uplift records derived from sea level records at 15 tide gauge stations along the northern

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